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Lithosphere dynamics and tectono-stratigraphic evolution of the Mesozoic Betic rifted margin (southeastern Spain)

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ABSTRACT

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Quantitative subsidence analysis demonstrates that Mesozoic sediments in the eastern Betic Cordilleras have been deposited on a thermally subsiding, rifted margin. The dynamics of rifting are related to regional trans-tension in the Atlantic realm. Modelling of this phase of rifting at the southern Iberian margin shows that the associated stretching of the lithosphere significantly altered its geometrical and mechanical structure and exerted a major control on subsequent convergence tectonics. The Mesozoic stretching led to the formation of a ramp at the margin due to strong lateral variations in the effective elastic thickness (EET) of the lithosphere and the thickness of the crust, in a direction perpendicular to the margin.

The existence of a ramp structure in the Late Mesozoic at the Iberian margin, which is also supported by stratigraphic and structural data, is important for the style of subsequent deformation of the lithosphere in Late Cretaceous–Tertiary times. The lateral variations in EET inferred from forward modelling of rifted margin stratigraphy could explain the observed multiple collapse of the Betic margin and the partitioning of deformation between the converging units involved in the Betic orogeny. Modelling of depth-dependent palaeo-rheological profiles of the margin shows that the mechanical properties of the lithosphere after stretching differed distinctly from those before stretching. Lithospheric deformation in the Betic Cordilleras during the Tertiary orogeny seems, therefore, to be largely controlled by the previous rifting phase.

A comparison of the palaeo and present-day rheologies in terms of estimates for EET during Late Cretaceous and present times indicates that the mechanical properties of the lithosphere underlying the Betic orogen were also affected by rifting during Late Oligocene–Aquitanian times. This phase of rifting is coeval with extension in the Valencia trough, the Gulf of Lions and the West European rift system, which supports a causal relation with regional lithospheric processes. The lateral and spatial variations in thermo-mechanical properties of the lithosphere underlying the Betics, therefore, reflect the dynamics of lithospheric extension on a larger scale.

Introduction

Many mountain belts have experienced polyphase tectonism, as they frequently represent former zones of rifting that are deformed and destroyed during a subsequent phase of convergence. The response of the lithosphere to vertical loading and horizontal stress in the convergence or orogenic stage is determined by the strength inherited from the foregoing rifting stage (e.g., Stockmal et al., 1986; Zoetemeijer et al., 1990). Therefore, better understanding of the lithospheric dynamics and basin development of a margin in collision will benefit from quantitative analyses of the preceding extensional history of the lithosphere.

We have performed such an investigation for the eastern Betic Cordilleras in southern Spain, which represent a Mesozoic rifted margin (García Hernandez et al., 1980) that has undergone major compressive deformation in the Tertiary. On tectono-stratigraphic grounds, the Cordilleras are subdivided into three zones (cf. Hermes, 1978; Fig. 1). These are from the north to the south: the Prebetic zone, the Subbetic zone and the Betic or Internal zone, respectively. The Subbetic

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zone is subdivided in two units: the North and the South Subbetic. The Prebetic zone corresponds to the proximal part of the southeastern Mesozoic Iberian passive margin, with sedimentary sequences characterized by shallow-water facies. The Subbetic deposits are of shallow-water facies in the Triassic–Early Jurassic and of pelagic facies from the Early Jurassic on, and represent the distal equivalent of the Prebetic (García Hernandez et al., 1980). The Internal zone in palaeo-geographical sense neither is related to the Prebetic nor the Subbetic (e.g., De Jong, 1990). The contact of the Internal zone with the Prebetic and the Subbetic is marked by the North Betic fault, which is a large-scale, high-angle fault (LeBlanc and Olivier, 1984). The contacts between the other units have a tectonic character as well. The Subbetic and the Prebetic are separated by a thrust zone, which has accommodated about 30 km displacement (Foucault, 1971), and as pointed out by Hermes (1978) and De Smet (1984), the contact of the North Subbetic and the South Subbetic is formed by the Crevillente fault zone (see also Fig. 2).

In this paper, we present the results of a quantitative subsidence analysis and stratigraphic modelling of the Prebetic and the Subbetic as well as synthetic palaeo-rheological profiles of the pre-orogenic Iberian margin. The purpose of this analysis is to quantify first-order features of the temporal and spatial variations in the dynamics of tectonic processes at the Mesozoic rifted margin of the southern part of the Iberian Peninsula. We also discuss the effect of Mesozoic stretching of the margin on flexural foreland basin development in the Tertiary. Finally we argue that the lithosphere underlying the Betics probably was affected by the same tensile stress regime that caused extension in the Valencia trough and other parts of the West European rift system.

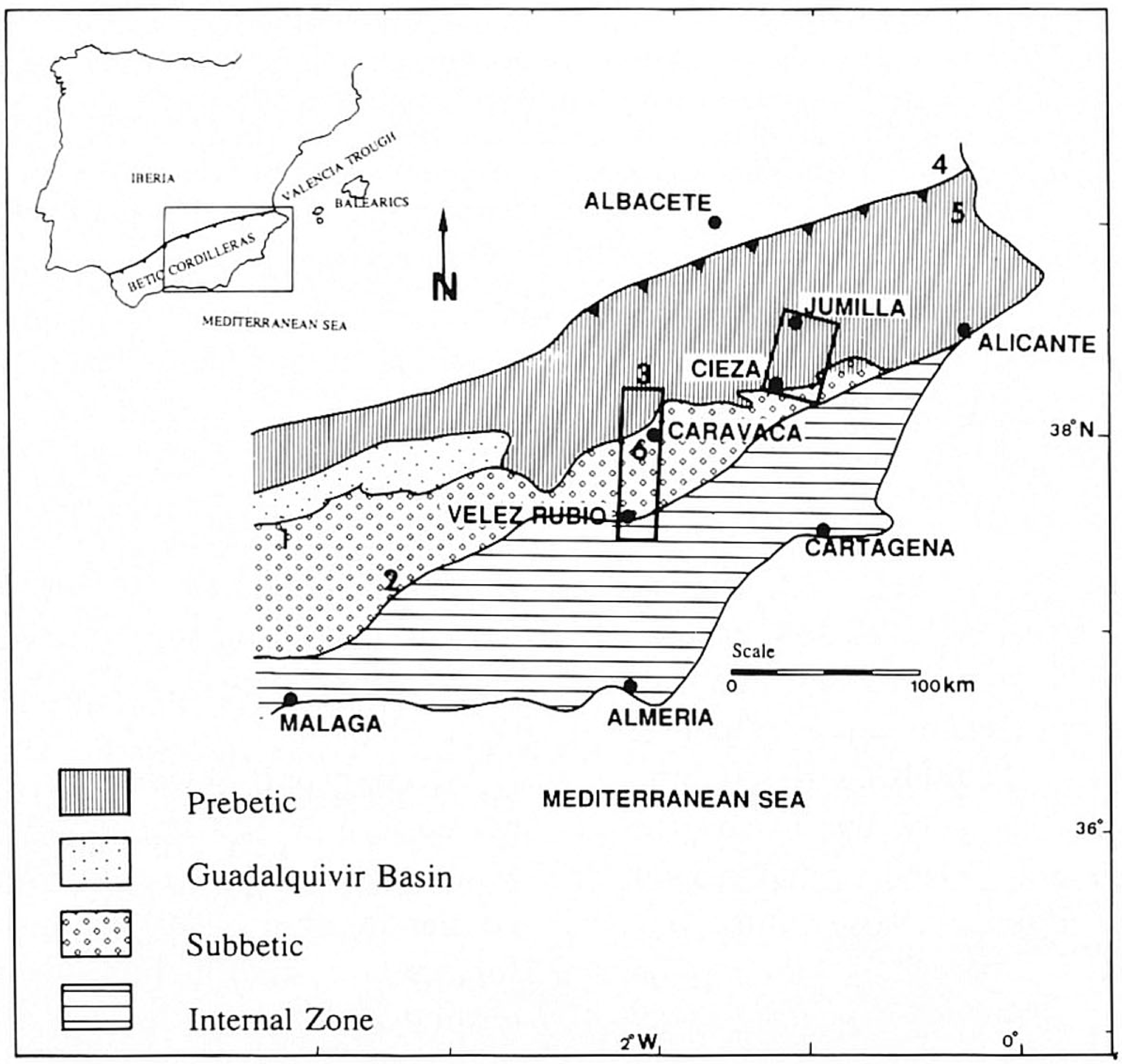

Fig. 1. Schematic geological map of southeastern Spain. Numbers mark the locations of analysed land sections and bore holes (see text and Lanaja, 1987). Bore holes: *1* = Nueva Carteya 1; *2* = Fuensante de Martos 1; *3* = Socovos 2; *4* = Jaraco 1; *5* = Perenchiza 1. Land section number: *6* = land sections situated on traverse marked by box (see also Fig. 2). Figure modified after Lanaja (1987). Inset shows geographical position. Boxes mark locations of investigated areas.

Subsidence analysis and tectonic evolution of the rifted margin

In the subsidence analysis, the tectonic component of the vertical movements of the basement is reconstructed, using backstripping techniques as discussed by Bond and Kominz (1984). We calculate basement and water loaded tectonic subsidence from the stratigraphic record, making corrections for compaction effects. For these corrections we use maximum and minimum porosity–depth relations for carbonate rocks, that cover the maximum and minimum porosities of the dominant lithologies in the analysed sequences (see also Bond and Kominz, 1984). Errors in the subsidence due to this approach are in the order of tens of metres and, as will appear, do not significantly affect the interpretation of the results. Water-depth variations in time, based on sedimentary facies and faunal contents of the rock, are incorporated. In the backstripping procedure, we did not assign a flexural strength to the basement underlying the load, and we ignored the blanketing effect of sediments. Corrections for these effects, however, also would not significantly affect the overall shape of the inferred tectonic subsidence curves (Watts et al., 1982; Lucazeau and Le Douaran, 1985). We did not correct the subsidence record for sea-level fluctuations (e.g., Vail et al., 1977). The eustatic components of the record of relative sea-level fluctuations are difficult to quantify and much of the sea-level record may actually be controlled by tectonic vertical motions (e.g., Cloetingh et al., 1985; Embry, 1990). In the absence of accurate data on the vertical motions during the time-spans covered by hiatuses in the sequences, we prefer to limit the number of assumptions in the analyses to a minimum.

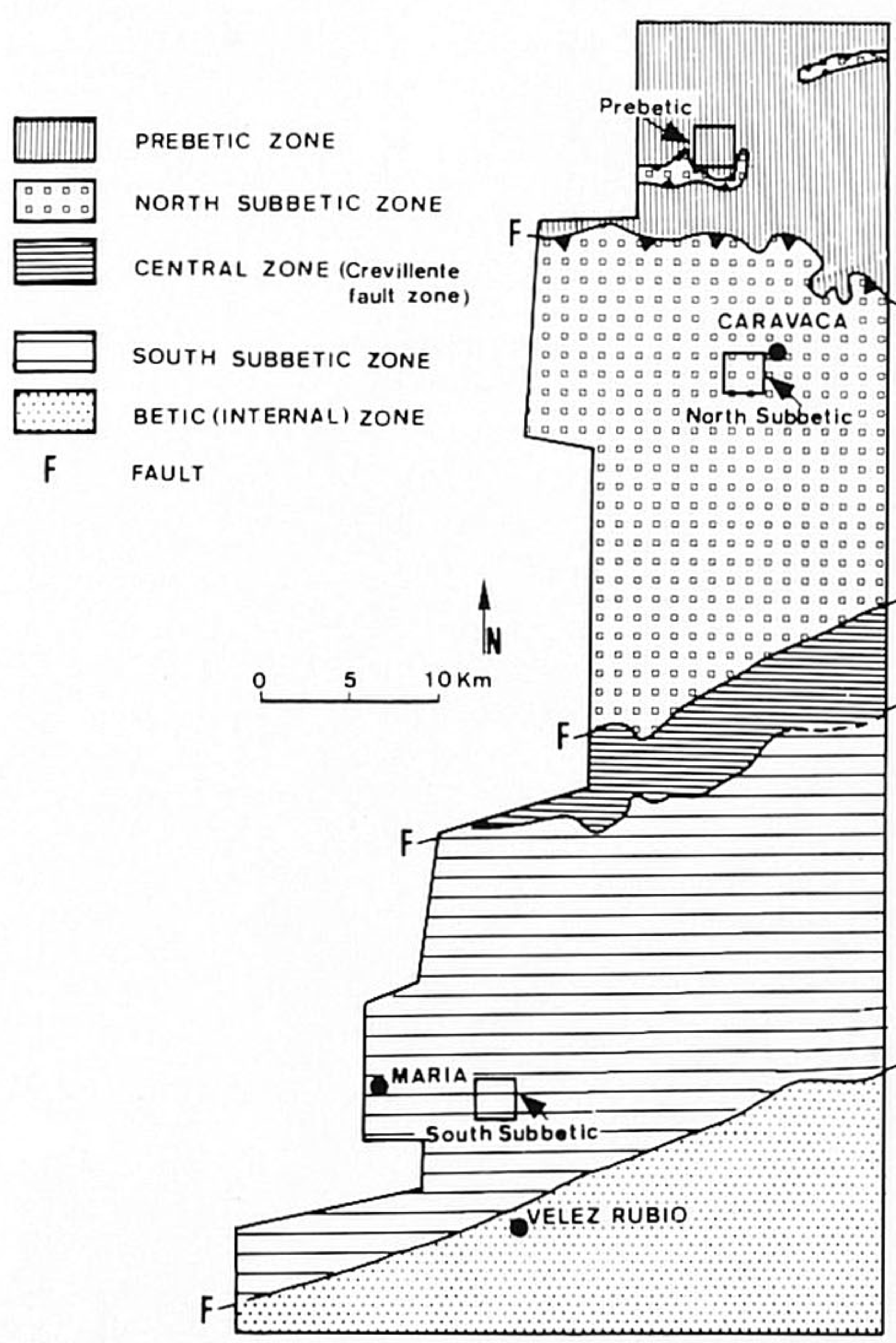

Fig. 2. Schematic geological map of the Velez Blanco–Caravaca area. Arrows mark locations of the analysed sections.

Stratigraphy

We have analysed stratigraphic sequences of the Prebetic, the Northern and the Southern Subbetic (cf. Hermes, 1978) in the area around Caravaca (Figs. 2 and 3). These sequences are well documented and the composition and the age of the rocks have been described extensively (Van Veen, 1969; Geel, 1973, 1979; Hoedemaecker, 1973; Hermes, 1978). Depositional depths are estimated from the sedimentary facies and faunal contents (T. Geel, pers. commun., 1989; Geel, 1973, 1979) while ages of datum levels are adopted from the Harland et al. (1982) time scale.

Although the Triassic is not exposed in the South Subbetic (Hermes, 1978), the Triassic probably was uniformly deposited over large areas (e.g., Besems and Simon, 1982; Simon, 1987). We, therefore, used the same thickness and composition of Triassic strata throughout. The stratigraphic subdivision of the Triassic, proposed by Paquet (1969) is modified after Besems and Simon (1982).

Composite well sections, catalogued by Lanaja (1987, fig. 1) have also been investigated. These sections bear no information on water depths, but tectonic signals still seem to be recognizable from the subsidence curves.

Subsidence of the rifted margin and regional tectonics

The tectonic subsidence curves for the North and South Subbetic show three phases of relatively rapid subsidence in the Mesozoic (Fig. 4). Conspicuously rapid subsidence is observed in the Middle Triassic, the Early Jurassic and the Callovian–Hauterivian. A comparison with subsidence patterns of the composite well sections (Fig. 5) shows that despite the uncertainties in the well data, a remarkable correlation occurs in the timing of increased subsidence in most of the wells of the southeastern Iberian plate and the land sections. This strongly suggests that the increased subsidence rates reflect large-scale tectonic processes.

During Mesozoic times, the plate motions of Iberia were closely related to reorganizations in the Atlantic–western Tethys domain (Ziegler, 1990). The overall stress regime in this domain during Mesozoic times was characterized by extension, and the phases of rapid subsidence coincide with Atlantic trans-tensional tectonics. The Triassic rifting coincides with distensive wrench tectonics throughout the Atlantic realm (Ziegler, 1988, 1990), the Early Jurassic rifting is coeval with early Kimmerian Atlantic rifting (Ziegler, 1988, 1990; Hiscott et al., 1990), and the Middle Jurassic–Early Cretaceous rifting is contempora-

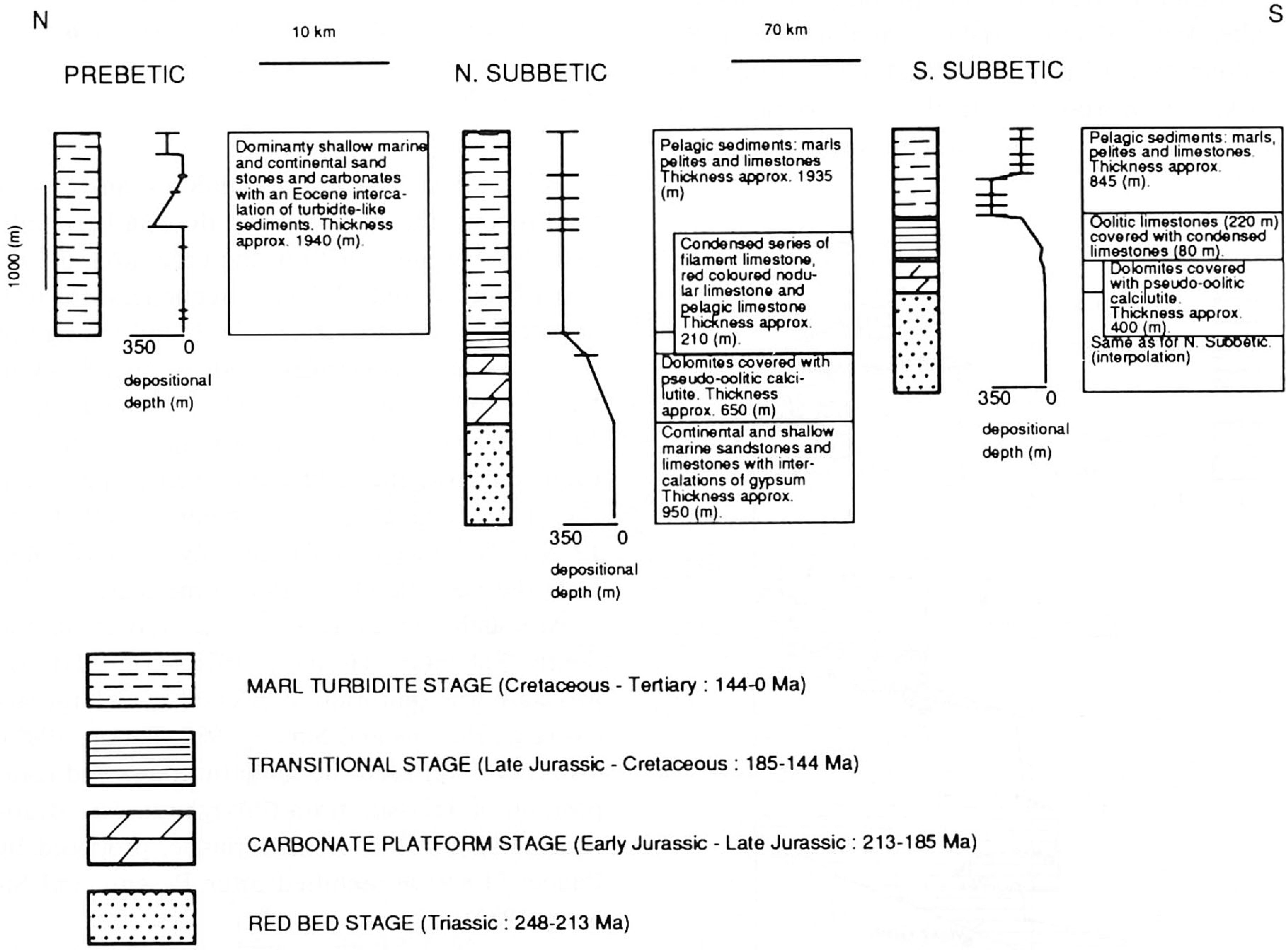

Fig. 3. Stratigraphic sections, with palaeo-water depths, of the Prebetic, the North Subbetic and the South Subbetic. Sections are aligned approximately from the north to the south in the present-day configuration. The annotations in the sections mark four stages that are distinguished in the stratigraphy of the Caravaca–Velez Blanco area: the Marl–Turbidite stage, the Transitional stage, the Carbonate Platform stage, and the Red Bed stage. Bars in the water-depth curves mark the error margins. (Based on data from Paquet, 1969; Van Veen, 1969; Hoedemaecker, 1973; T. Geel, pers. commun., 1973, 1979; Hermes, 1978; Simon, 1987.)

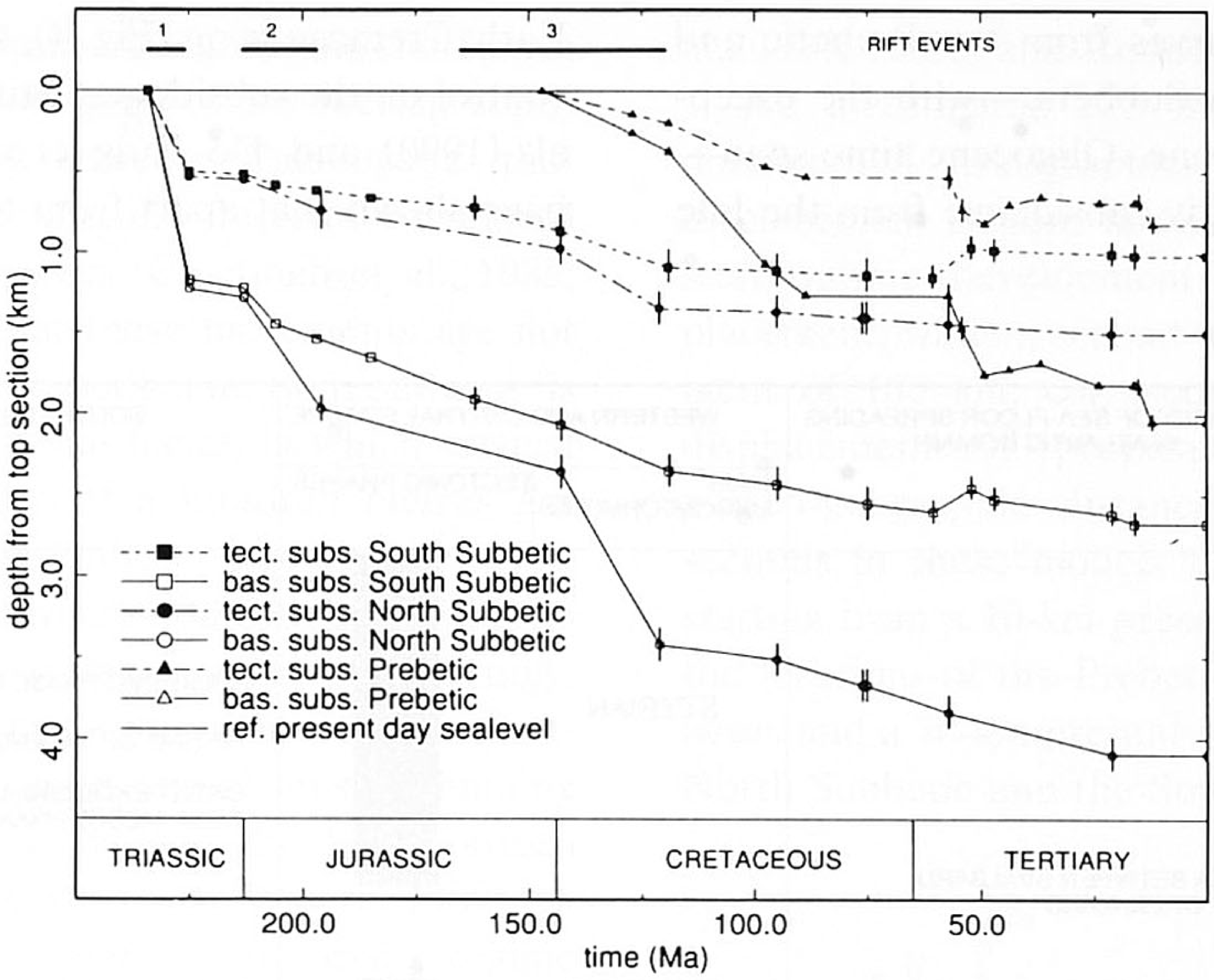

Fig. 4. Basement and tectonic subsidence curves from the Prebetic, the North and the South Subbetic. Continuous lines are basement subsidence curves, dashed lines are tectonic subsidence curves. Vertical bars represent errors in the estimates for palaeo-bathymetry. Horizontal bars marked by numbers 1–3 indicate the timing and duration of subsequent rifting phases at the Betic margin.

neous with the initiation of the Central and South Atlantic rift (Savostin et al., 1986; Ziegler, 1988, 1990; see also Fig. 6). Therefore, it seems that the observed phases of increased subsidence reflect regional rifting. From the subsidence curves we estimate the approximate duration of the subsequent rifting events in the Caravaca area to be 8 Ma, 16 Ma and 45 Ma respectively.

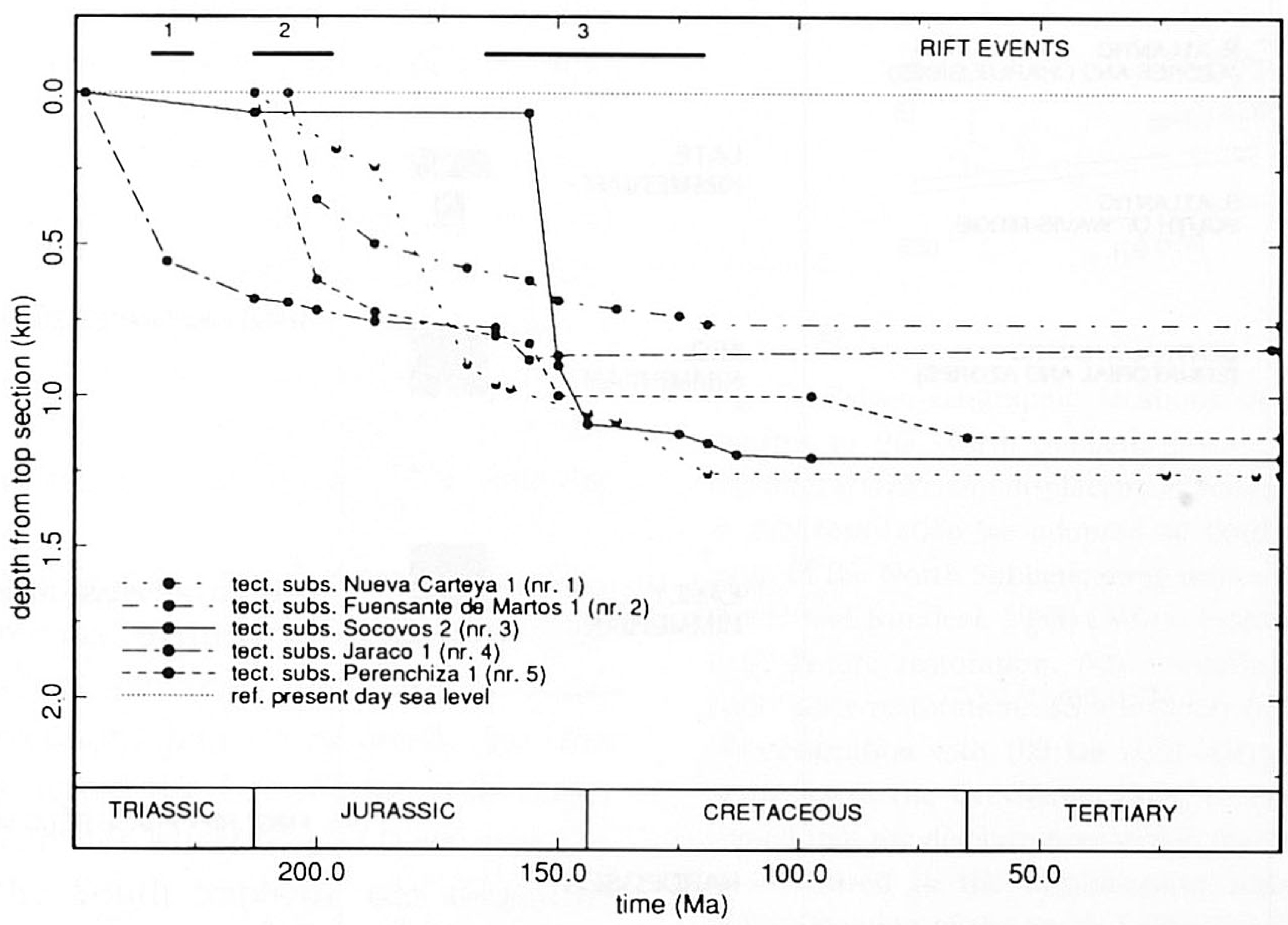

Fig. 5. Tectonic subsidence curves from composite well sections across the Betic margin. The locations of the bore holes are marked on the map in Fig. 1. Horizontal bars marked by numbers 1–3 indicate the timing and duration of subsequent rifting phases at the Betic margin.

The subsidence curves from the Prebetic and the North and South Subbetic—with the exception of the Late Eocene–Oligocene time-span—generally show a steady subsidence from the late Early Cretaceous on (Fig. 4), suggesting a thermal control on the subsidence. Studies from Kenter et al. (1990) and De Ruig et al. (1991), however, have shown that apart from these long-term sub-

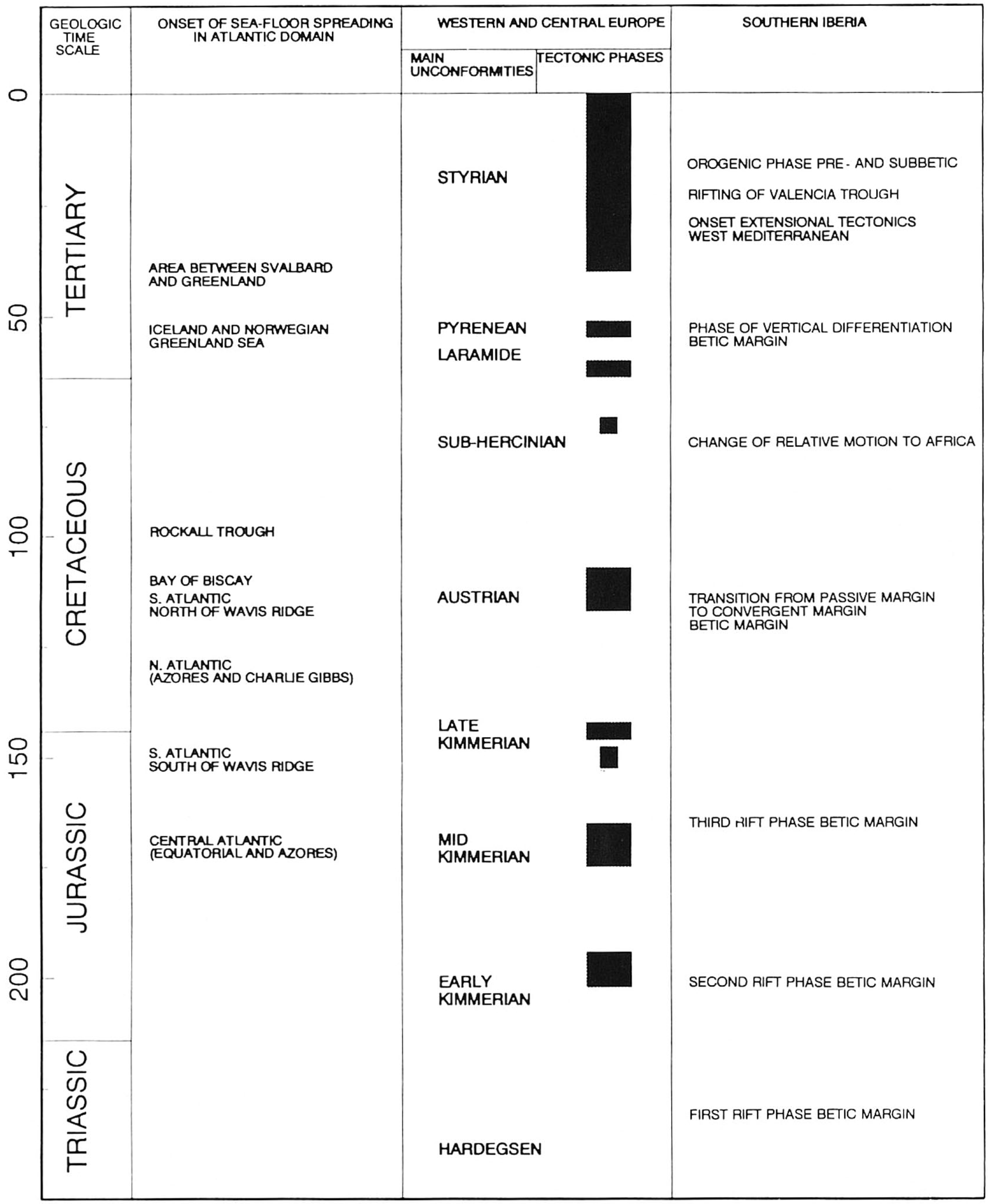

Fig. 6. Schematic display of timing of regional tectonic events in the West European–Atlantic domain and comparison with tectonic phases in southern Iberia. Figure modified after Ziegler (1990).

sidence patterns, there are significant short-term vertical motions of the shelf of the Iberian continent in the Late Cretaceous–Palaeogene, that are related to far-field stress fluctuations related to convergence processes (Cloetingh et al., 1985; Cloetingh, 1988). That these movements are not reflected in Subbetic sedimentation patterns is probably due to its distal facies, in which vertical motions of the order of a hundred metres are beyond the detection limit.

The Tertiary geodynamic history of the Betics is not characterized by convergence tectonics only. During the Oligocene–Aquitanian period, extensional tectonics caused renewed crustal thinning of the lithosphere underlying the Betic orogen (Bakker et al., 1989; Platt and Vissers, 1989). In the Burdigalian–Tortonian, a compressive regime is generated again, whereby compression induced the reactivation and generation of thrust faults and wrench zones (Montenat et al., 1987; Nobel and Rondeel, 1988), and initiated the last orogenic phase of the Betic fold-and-thrust belt.

Quantitative modelling of the pre-orogenic stratigraphy

The pre-orogenic stratigraphy of the Iberian margin reflects the extensional history of the margin, which controls the evolution of the thermal–mechanical structure of the underlying lithosphere. The rheological properties of the continental lithosphere largely depend on crustal thickness and on the thermal structure of the lithosphere (Kirby, 1983). To determine a palaeo-rheological profile of the Iberian margin, we have carried out a forward modelling of the pre-orogenic stratigraphy of the Prebetic and the Subbetic.

Modelling of the Late Mesozoic stratigraphy requires the reconstruction of the palaeo-geographical distances between the land sections. Therefore, corrections had to be made for the lateral motions across the Crevillente fault zone. According to Van der Fliert et al. (1980) and De Smet (1984), the South Subbetic was displaced 400 km right-laterally across the fault, while other authors suggest displacements from 0 up to 100 km (Sanz de Galdeano, 1983, 1988; De Ruig et al., 1987; Nobel and Rondeel, 1988).

We investigated two models to explain the documented stratigraphies. One model with a displacement of zero km was used to investigate stratigraphic development with minimum displacement, while a second model with a displacement of 100 km, corresponds to the maximum displacement of proposed alternatives. The palaeo-geographic distances between the land sections in these models have been determined starting from a 10-km present-day offset between the locations of the Prebetic and the North Subbetic, and a 70-km present-day offset between the North Subbetic and the South Subbetic. Further-

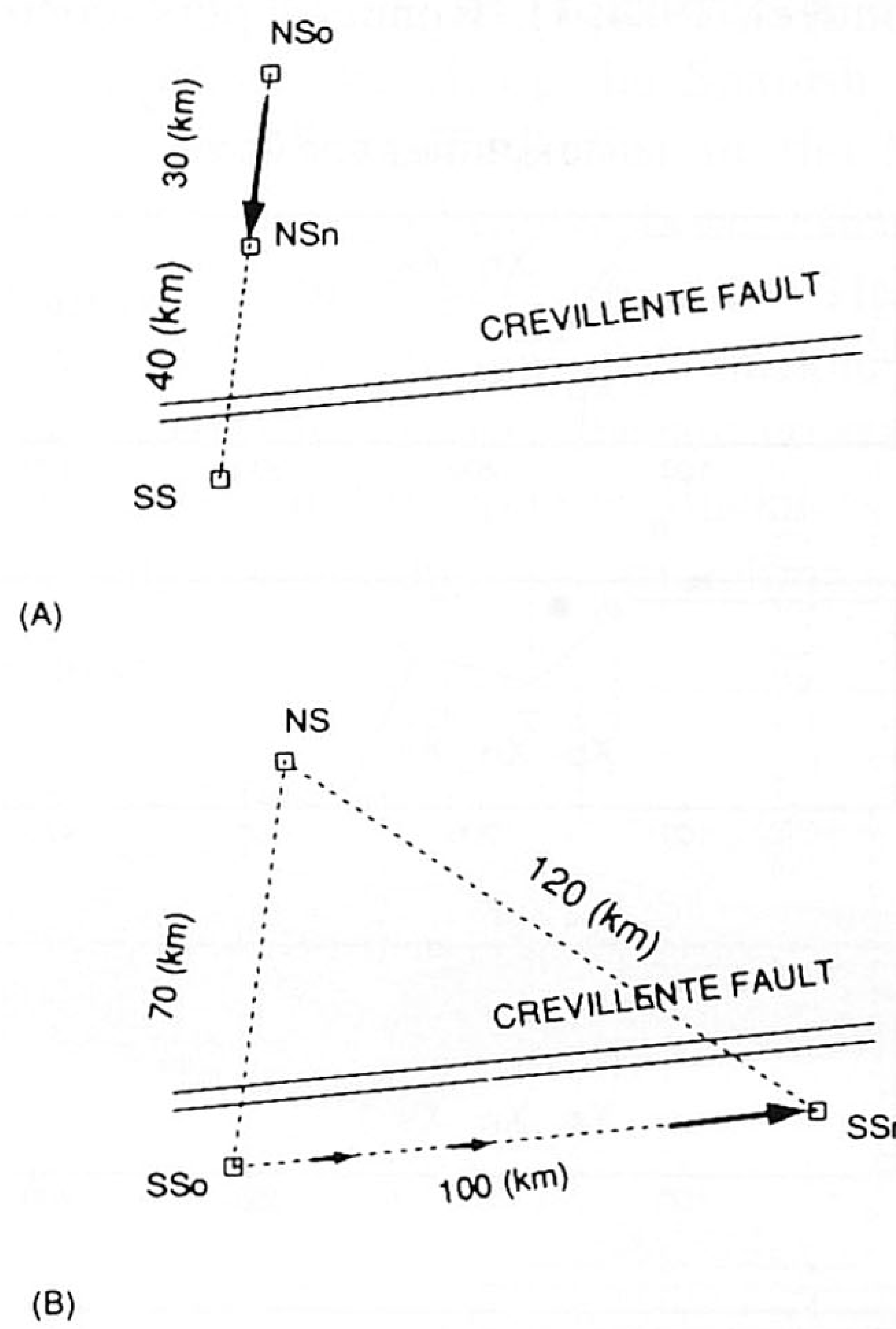

Fig. 7. Palaeo-geographic locations of the South Subbetic relative to the North Subbetic restored with: (A) zero km left-lateral strike-slip displacement across the Crevillente fault; in this restoration we adopted 40 (km) orthogonal displacement of the North Subbetic away from the South Subbetic (cf. Nobel and Rondeel, 1988) (*NSo* = location of the North Subbetic before restoration, *NSn* = location of the North Subbetic after restoration, *SS* = location of the South Subbetic; (B) restoration with 100 km right-lateral strike-slip displacement across the Crevillente fault; in this restoration we assumed that the displacement within the separate units is small as compared to the displacement across the major faults (*SSo* = location of the South Subbetic before restoration, *SSn* = location of the South Subbetic after restoration, *NS* = location of the North Subbetic). Arrows mark the restoration to the original location.

more, we adopted a northward displacement of 30 km of the North Subbetic across the Prebetic (Foucault, 1971) and relatively minor shortening within the units themselves. Minor shortening was adopted, starting from the point that major displacement occurred along the main fault zones that separate the units. In the case of the 100-km displacement model, values of 120 km for the distance between the North Subbetic and the South Subbetic and 40 km for the distance between the North Subbetic and the Prebetic were obtained (Fig. 7A). Similarly, but now disregarding any horizontal displacement across the fault, and assuming that the North Subbetic orthogonally slid away from the South Subbetic (Nobel and Rondeel, 1988; H. Rondeel, pers. commun., 1990), values of 40 km were obtained for both the Prebetic and the North Subbetic and the palaeogeographical distance between the North Subbetic and the South Subbetic (Fig. 7B).

The forward modelling of the stratigraphy is based on a mechanism of stretching of an elastic plate, using β (extension factor) estimates inferred from the subsidence analysis. Thinning of the plate is calculated starting from a pure-shear model (McKenzie, 1978). Simple-shear models (Wernicke, 1985) require information on the geometry of major pre-Oligocene intraplate detachment zones, which currently is not available for this part of the Betics. Therefore, components of simple shear have been ignored in the present study. We will show that the Mesozoic tectonics

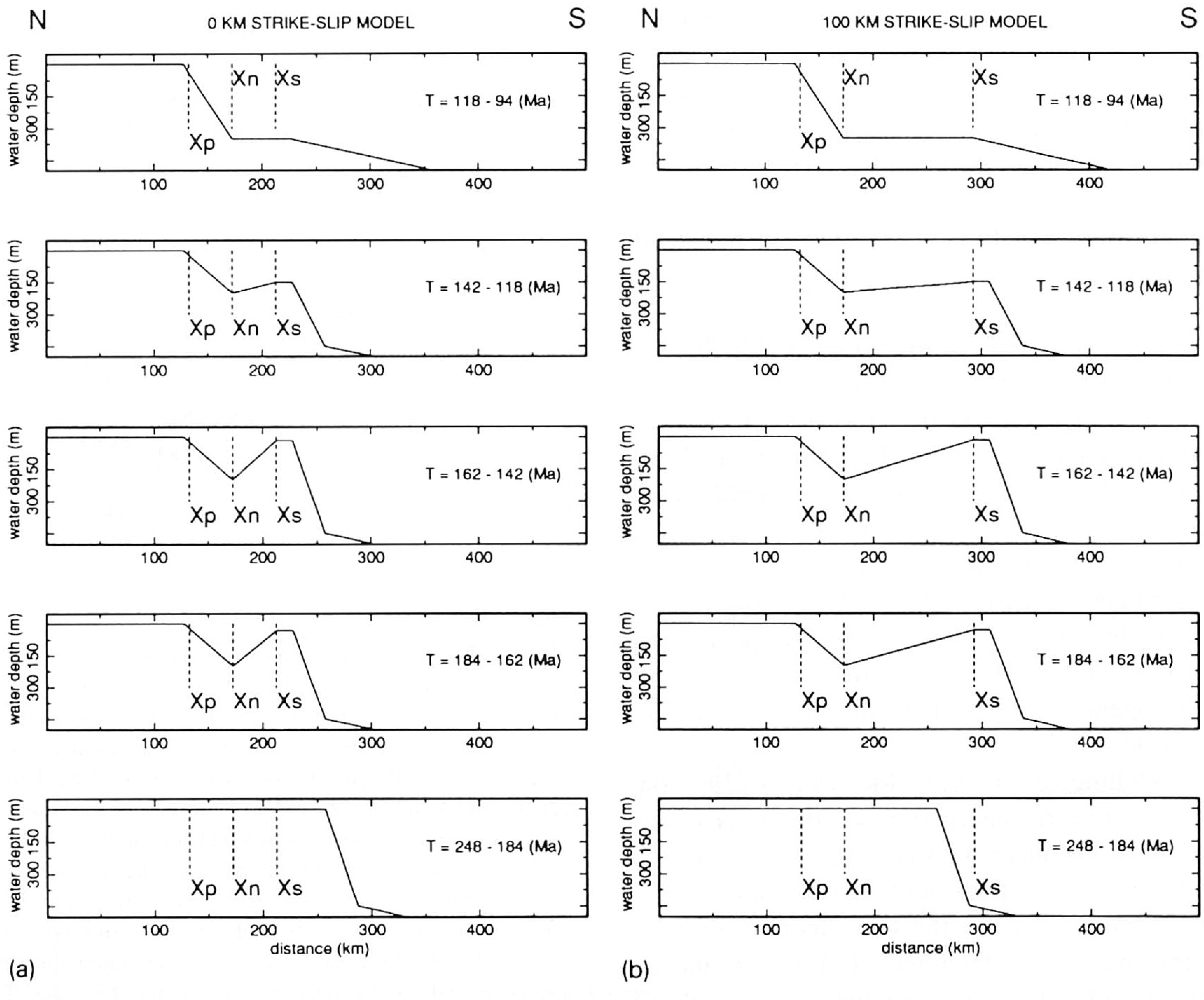

Fig. 8. Synthetic palaeo-bathymetry profiles for the Iberian margin as a function of time. Figures represent palaeo-bathymetry for (A) a zero km strike-slip displacement model and (B) a 100-km strike-slip displacement model for time intervals (from bottom to top) 248–184 Ma, 184–162 Ma, 162–142 Ma, 142–118 Ma, and 118–94 Ma). X_p, X_n and X_s represent the Mesozoic locations of the Prebetic, the North Subbetic and the South Subbetic sections, respectively (see also text).

of the Betic margin can largely be explained with a model incorporating only pure shear.

The vertical space created as a result of crustal thinning and subsequent thermal subsidence (cf. Sclater and Christie, 1980; Steckler et al., 1988) is filled with sediments of density 2.6 g/cm^3 (Thorne and Watts, 1989). Water-depth profiles used in the model are reconstructed from the sediment characteristics in the land sections (T. Geel, pers. commun., 1989; Geel, 1973, 1979; see also Fig. 8) and are reasonably well constrained in the areas between and surrounding the sections (see also García Hernandez et al., 1980). In the area north of the Prebetic section, a zero meter water depth is adopted, based on palaeo-geographic studies (e.g., García-Hernandez et al., 1980; Nobel and Rondeel, 1988). In the area within 50 km south of the South Subbetic, the water depths probably gradually increased (e.g., García Hernandez et al., 1980; T. Geel, pers. commun., 1989). For the area located more than 50 km south of the South Subbetic, we also adopted a smooth increase of water depths over a large distance. This assumption implies deposition of approximately equally thick packages of sediments over large distances, and was made to reduce the errors in the flexural solution due to the lack of information about this area.

The flexural solution is calculated with an effective elastic thickness (EET) corresponding to the depth to an isotherm. Changes of the EET due to heating with stretching, and cooling due to vertical and lateral heat conduction are calculated, using a finite difference approach. Models are developed for an initial EET of 23 km and of 30 km, respectively, which are moderate values for relatively undisturbed continental lithosphere (Watts et al., 1982; Thorne and Watts, 1989). The value of the initial crustal thickness is taken at 32 km, corresponding to the mean thickness of the crust at present underlying the Spanish Meseta (Banda et al., 1983). The crust of the Spanish Meseta has not significantly been affected by major tectonic processes after the Hercynian orogeny, and the present-day thickness thus roughly reflects the pre-rift, Iberian crustal thickness. Changes in lithospheric thickness in the model only occur due to stretching. Conse-

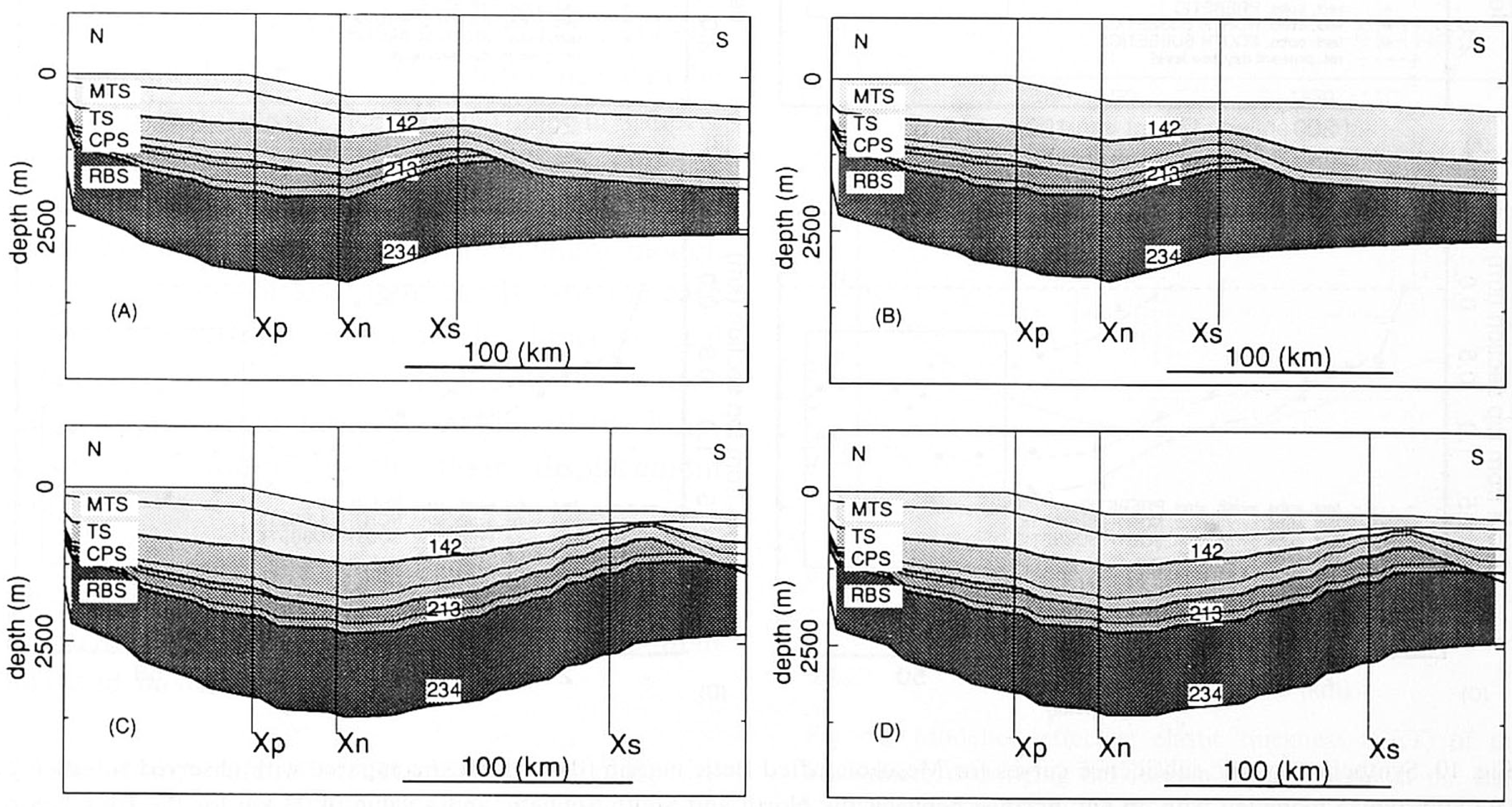

Fig. 9. Modelled stratigraphy for the Iberian rifted margin adopting multiple stretching phases. (A, B) The stratigraphy modelled with 40 km distance between the North and the South Subbetic and values of 23 and 30 km, respectively, for the EET before stretching. (C, D) The stratigraphy modelled with 120 km distance between the North and the South Subbetic and values of 23 and 30 km, respectively, for the EET before stretching. Abbreviations: *MTS* = Marl–Turbidite stage; *TS* = Transitional stage; *CPS* = Carbonate Platform stage; *RBS* = Red Bed stage. Numbers mark ages of underlying horizons.

quences of blanketing effects on lithospheric cooling have been ignored, in the absence of a large set of data on the conductivity parameters of the studied sediments.

Figure 9 displays the modelled stratigraphies. The figure shows some truncation patterns in the stratigraphies. In the absence of accurate information on the true Mesozoic bathymetrical profile, the existence of these truncations at present cannot be sustained with geological data. The predicted truncations reflect the adopted depositional depths in the models, and thus, considering the intrinsic uncertainties in the palaeo-bathymetry for larger water depths (see also Fig. 3), may in some cases be artefacts of the calculations. However, since the area of erosion is extensive, whereby the amount of erosion in between the sections and the area surrounding the sections is relatively small, the uncertainties in the calculations do not significantly affect our interpretation of the modelling results.

The predictions of the synthetic stratigraphies can be checked by comparison of subsidence of the synthetic sections and the observed sections of the Prebetic, the North Subbetic and the South Subbetic (Fig. 10). The β values, that are adopted to obtain a fit for the distinct models, are different (see also Fig. 11). The thermal and deforma-

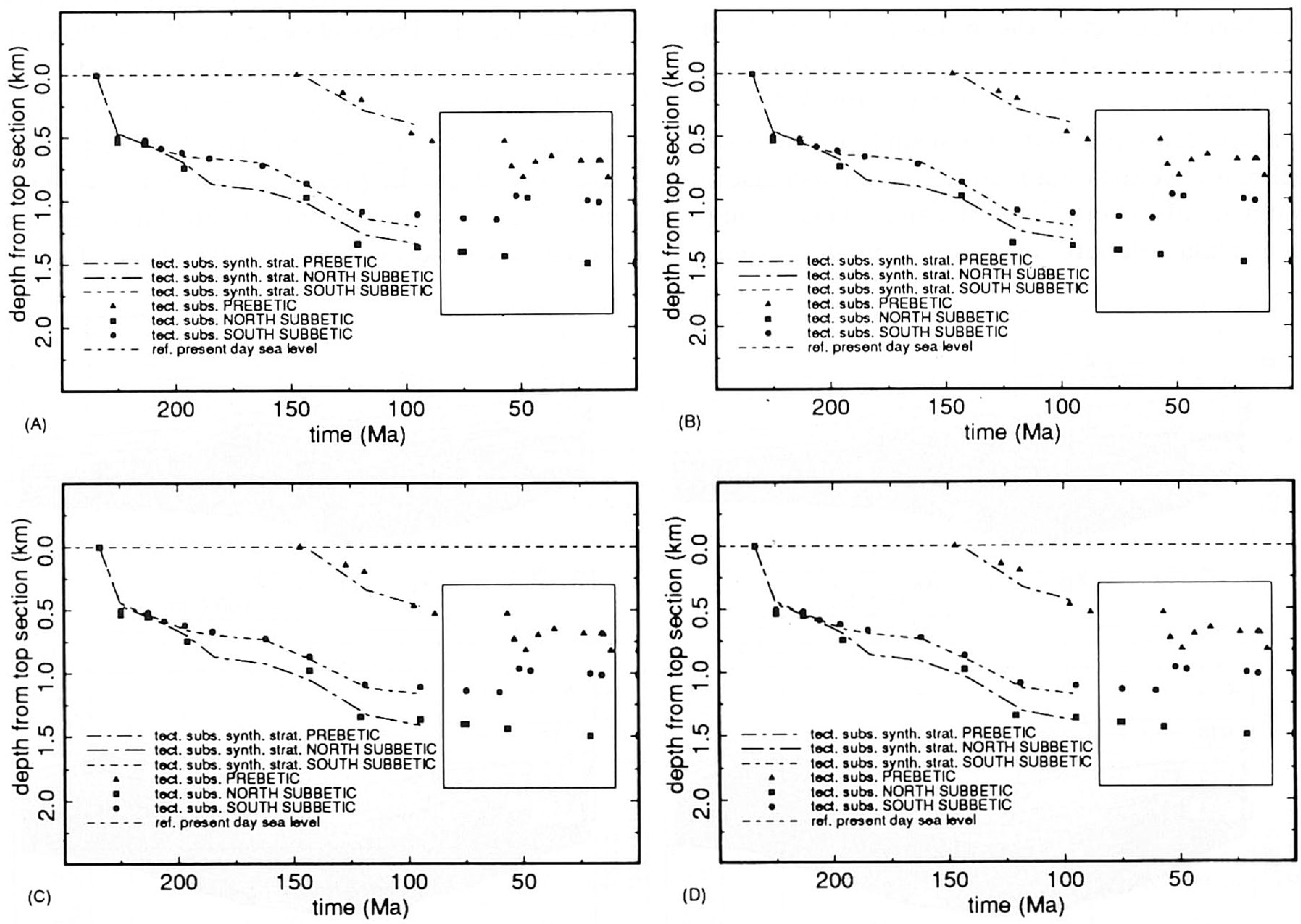

Fig. 10. Synthetic tectonic subsidence curves for Mesozoic rifted Betic margin (dashed lines) compared with observed subsidence. (A) Subsidence modelled with 40 km distance between the North and South Subbetic and a value of 23 km for the EET before stretching. (B) Subsidence modelled with 40 km distance between the North and South Subbetic and a value of 30 km for the EET before stretching. (C) Subsidence modelled with 120 km distance between the North and South Subbetic and a value of 23 km for the EET before stretching. (D) Subsidence modelled with 120 km distance between the North and South Subbetic and a value of 30 km for the EET before stretching. Subsidence data in boxes mark the subsequent record of vertical motions, which is strongly affected by Tertiary post-rift tectonics.

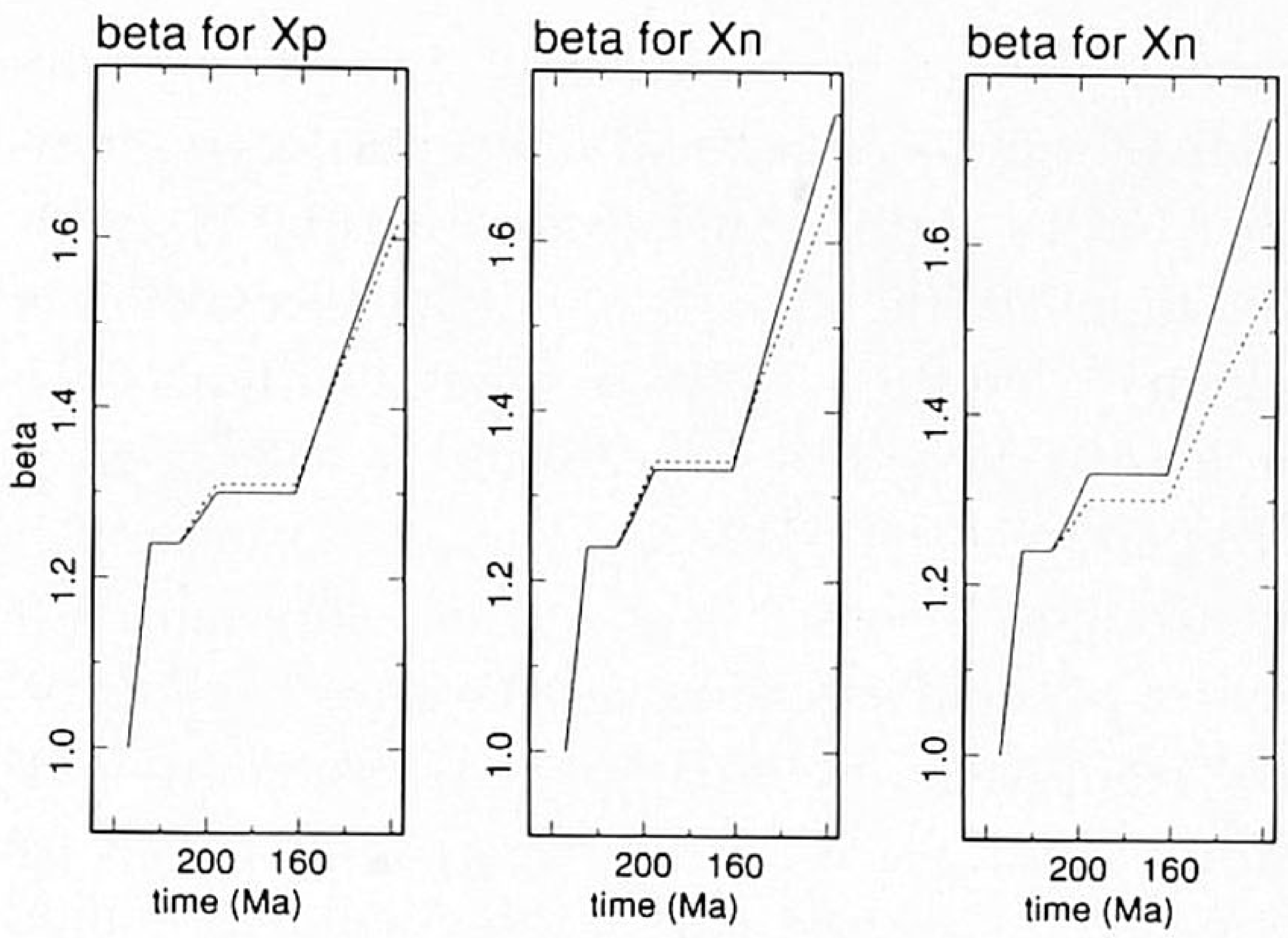

Fig. 11. Values of β, adopted at the locations of the land sections, to obtain fits for the 2-D stratigraphic modelling. β is represented as a function of time. Note β (x, t) varies for distinct models: dashed lines represent β for the model with 0 km displacement across the Crevillente fault, continuous lines represent β for the model with 100 km displacement across the Crevillente fault.

tion histories predicted by the models thus are different, and consequently the lithospheric strength evolved dissimilarly as a function of time.

The relative values of β for the 0-km displacement model and β for the 100-km displacement model differ for the Prebetic, the North Subbetic and the South Subbetic. The difference can be explained in terms of flexure of the lithosphere. Sediment-loading in the North Subbetic and the Prebetic probably causes flexural uplift of the South Subbetic in the 0-km displacement model, which may not—or less significantly—be the case in the 100-km displacement model. Since vertical subsidence of the South Subbetic in the models has to be the same, larger stretching of the lithosphere is required by the 0-km displacement model.

Mechanical structure of the rifted margin and implications for the transition to the subsequent foreland basin phase

The temperature structure of the plate inferred from the modelling of the synthetic stratigraphy of the overlying rift basin yields estimates for the EET as a function of space in the pre-orogenic Iberian continental lithosphere (Fig. 12). All models predict a large decrease of the value of the EET from north to south. For the model with an initial EET of 23 km, the decrease is 9 km over a lateral distance of 100 km. For the model with an initial EET of 30 km, the decrease is 8 km over a horizontal distance of 100 km. Models with a larger initial EET would require approximately similar β values as models with an initial EET of 23 and 30 km, respectively (cf. Steckler et al., 1988), which implies that the ratio of the pre-rift EET and post-rift EET in such models would approximately be the same with larger values of the initial EET.

Work by Stockmal et al. (1986) has emphasized the importance of lateral variation of the

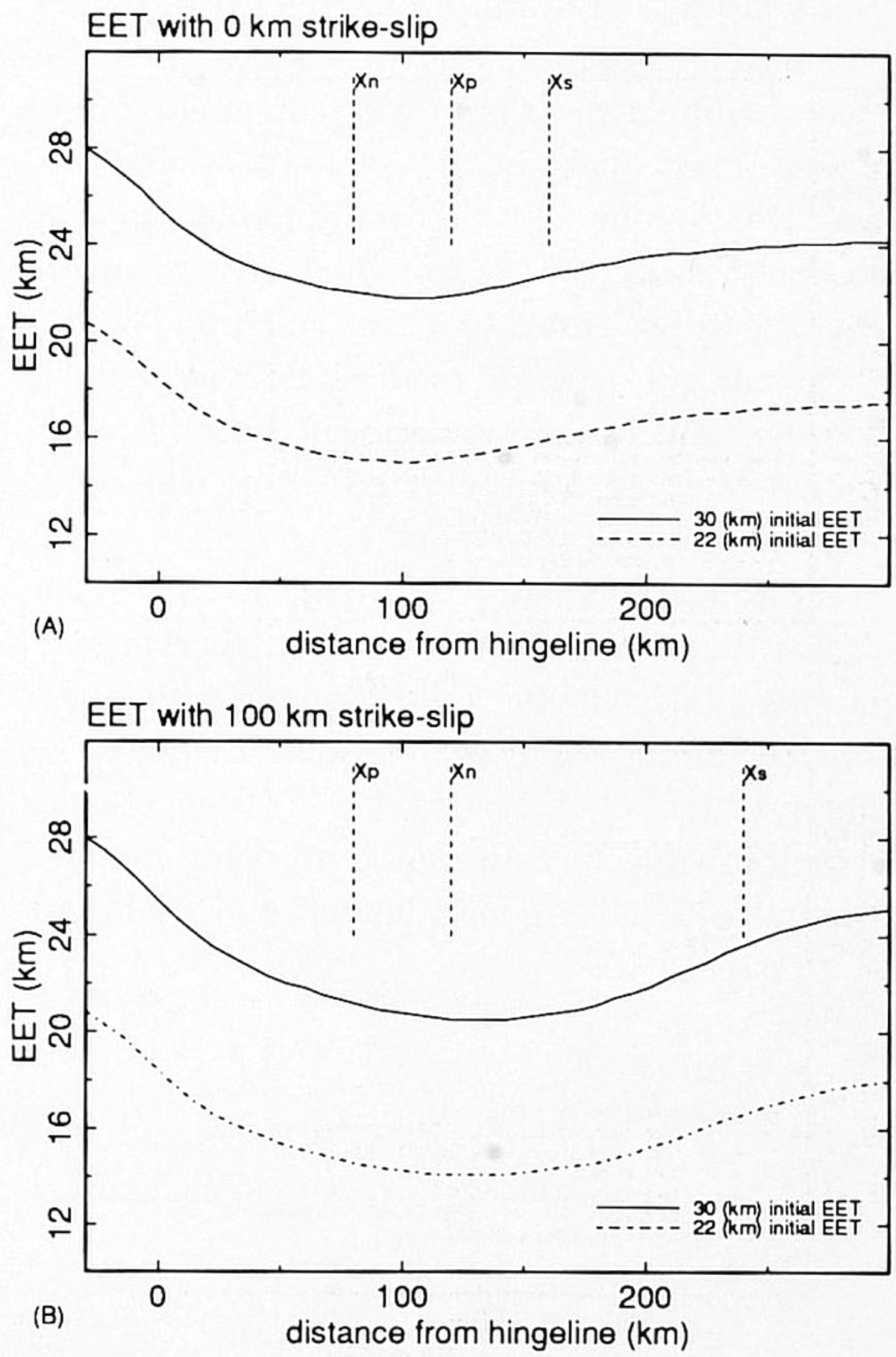

Fig. 12. Modelled effective elastic thickness (EET) of the lithosphere underlying the Iberian rifted margin in the Late Cretaceous. The EET is defined by the depth to an isotherm. The depth of the isotherm changed due to stretching. Models are developed for cases of 40 km (A) and 120 km (B) palaeogeographic distance between the North and South Subbetic with values of 23 (dashed lines) and 30 km (continuous lines) for the EET before stretching.

EET in the development of rift and foreland basins. These authors pointed out that horizontal and vertical loading of a previously rifted margin with a significant decrease of the thickness of the lithosphere underlying the margin, generates small deep basins, whereas loading in case of a more moderate decrease generates broader shallower basins (Fig. 13). Basins underlain by lithosphere with substantial lateral EET variations also differ strongly from basins underlain by lithosphere with a more moderate EET variation in their subsequent foreland basin evolution. This as a consequence of the formation of a ramp structure in the lithosphere when strong lateral variations in the EET occur over short distances. Overthrusting loads are pin-pointed at the ramp, causing major deformation to take place in the region south of the ramp, whereas minor deformation takes place in the more continentward area. Continuous shortening ultimately leads to thrusting across the ramp, which results in another phase of foreland basin collapse. With a less significant decrease of the EET, the tectonic processes during shortening will become more gradual and the impact of collapses will be of minor importance.

These features are important, as field data suggest that the Mesozoic southern Iberian margin was underlain by a lithosphere with major decrease of the EET over a short distance. As demonstrated by De Jong (1990), convergence tectonics in the Iberian region probably initiated at about 110 Ma. Angular unconformities in Aptian sediments (approximately 110 Ma) in the Prebetic near Cieza record tectonic reorganizations in this period (for location see Fig. 1). From this time on, the sedimentary sequences show a regional deviation from a simple pattern of a prolonged period of consistently unchanging facies patterns. In the Cieza–Pila area for example, relatively long phases of quiescence alternate with phases of rapid submergence (Droste, 1984). Similar events are observed in the surrounding area (Kenter et al., 1990). Field data also indicate that deformation in the Internal zone, which probably formed a convergent unit during the Late Cretaceous–Palaeogene (e.g., De Jong, 1990), is much larger than in the adjacent areas (Egeler and Simon, 1969). Thus, the southern Iberian margin probably was underlain by a laterally rapid thinning lithosphere. Therefore, the Betic orogenic basin development is probably strongly controlled by the preceding rifting stage.

From the predicted temperature distribution of the lithosphere and the calculated thinning of the crust underlying the rifted margin, we have calculated palaeo-rheological profiles of the Iberian continental lithosphere (Fig. 14). To construct palaeo-rheologies (see also Ranalli and Murphy, 1987) we have adopted a three-layer petrological model for the lithosphere, with an upper crustal layer of wet quartz, a lower crustal layer of wet diorite and a subcrustal layer of olivine (cf. Brace and Kohlstedt, 1980; Kirby, 1983; Carter and Tsenn, 1987). Since there is no current consensus about the mechanism of thin-

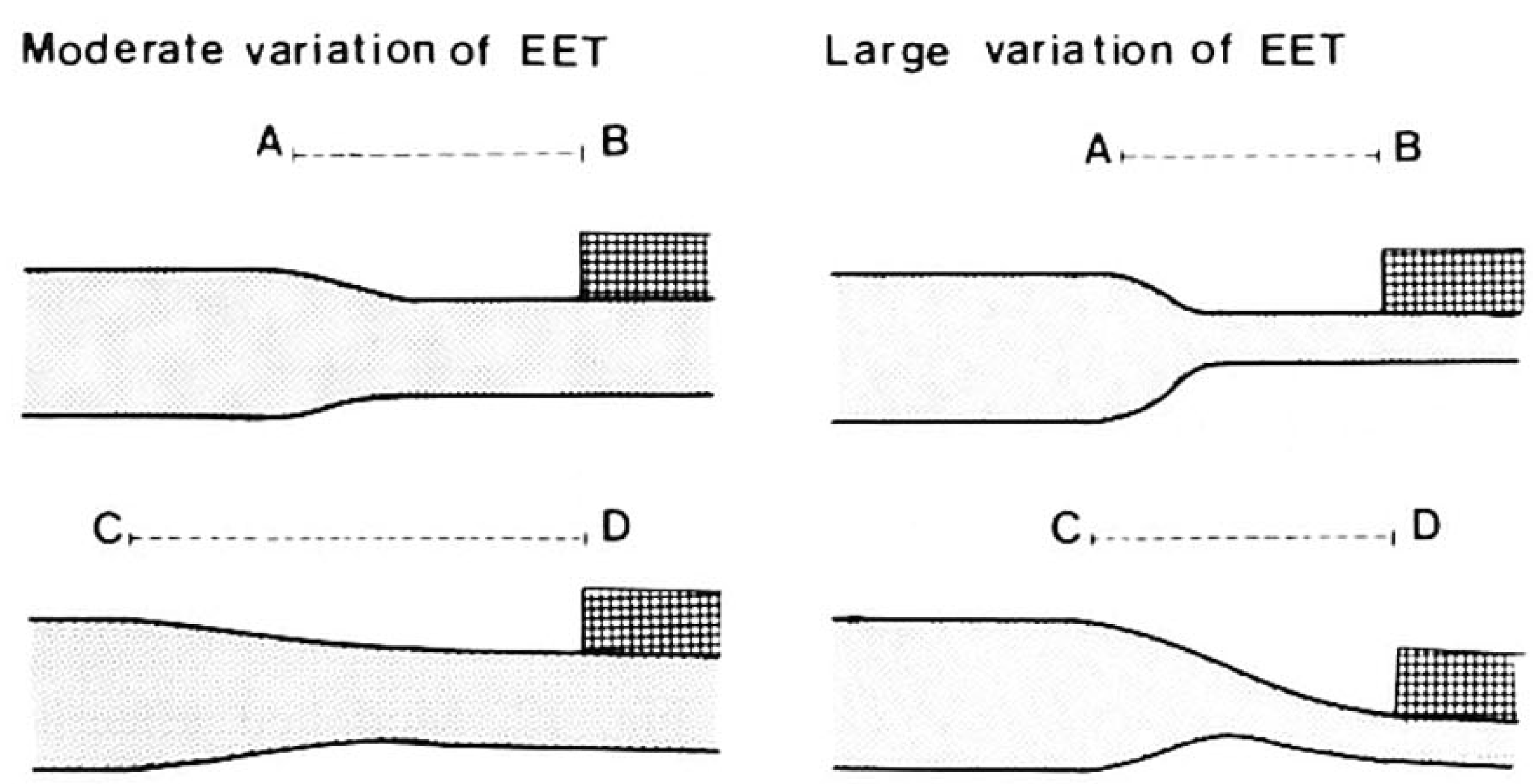

Fig. 13. Effect of a lateral decrease of the EET of the rifted lithosphere on subsequent foreland basin development. The upper panel shows vertically loaded margins with different EET. The intervals *A–B* mark the distance between the load and the edge of the continental area. The lower panel shows the margins after deflection, in which the intervals *C–D* mark the distance between the load and the continental area. A moderate lateral decrease of the EET leads to the formation of shallow, broad basins in front of the load; a more rapid lateral variation in EET leads to the formation of deeper, narrower basins. Figure modified after Stockmal et al. (1986).

ning of the lower crust versus thinning of the upper crust (e.g., Pinet and Bois, 1989), we have adopted equal amounts of thinning of the upper and the lower crustal layer, limiting the number of assumptions to a minimum. The values of the strain rates and the densities of the rocks used in the calculation of the strength, are given in Table 1. As to investigate the lateral resolution of the profiles, we have carried out numerical experiments with different stretching parameters for the crust. These calculations show that the significance of these parameters, and thus the lateral resolution of the rheological profiles, pertains to approximately 35 km from the stratigraphic sections. The results for the models with similar displacement but with different initial EET are nearly identical. Therefore, we only present two models with different displacement.

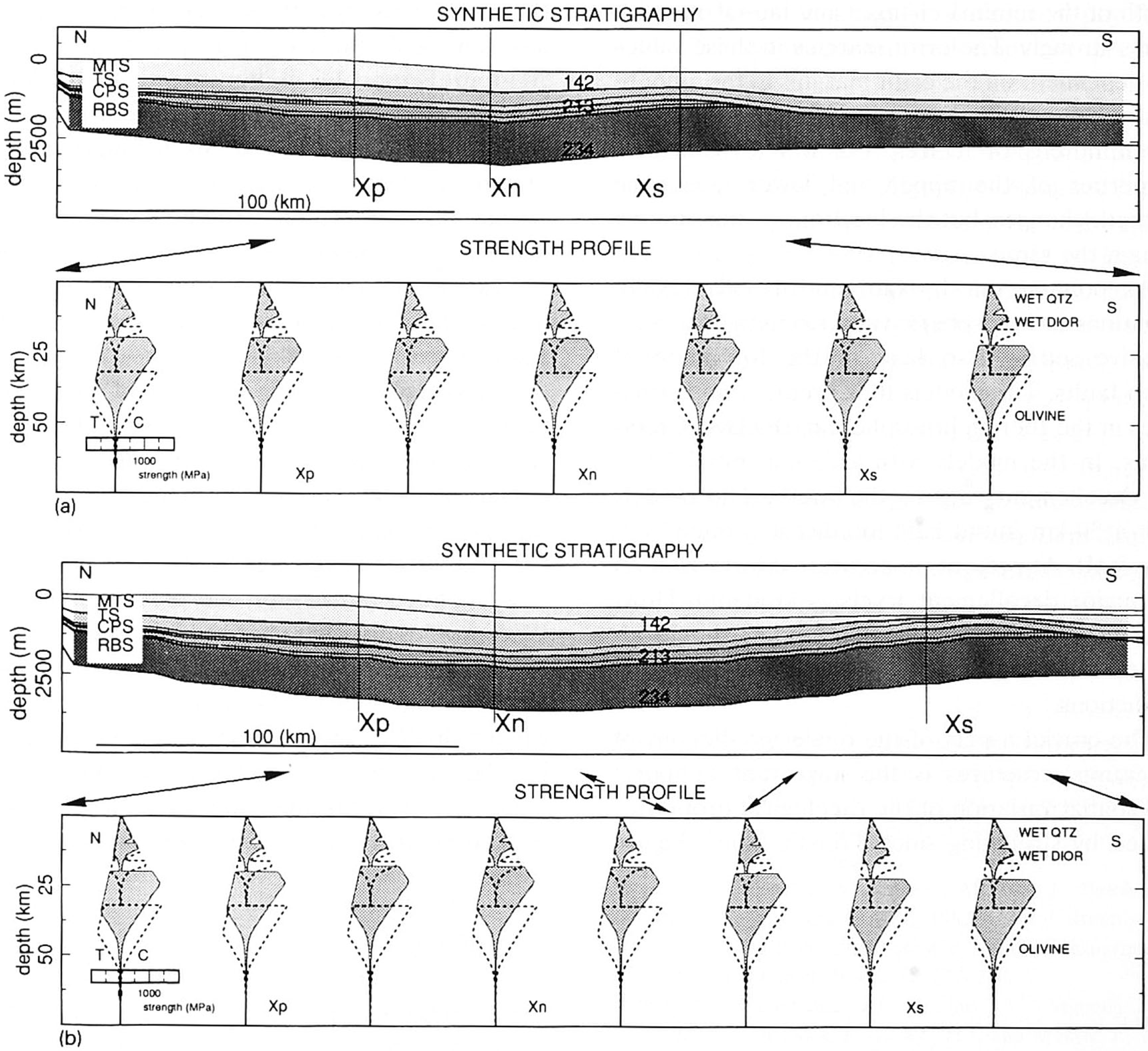

Fig. 14. Palaeo-rheological profiles of the Triassic and the Late Cretaceous Iberian margin. The profiles display depth-dependent yield envelopes, calculated with a three-layer model for the lithosphere, whereby the lithosphere consists of an upper crustal wet quartz layer, a lower crustal wet diorite layer and an underlying olivine layer. Dashed envelopes are Triassic palaeo-rheological profiles; continuous, filled envelopes are Late Cretaceous strength profiles. (A) Model with 40 km palaeo-geographic distance between the North and the South Subbetic, and a value of 30 km for the EET before stretching. (B) Model with 120 km palaeo-geographic distance between the North and South Subbetic, again with a value of 30 km for the EET before stretching. Arrows connecting the upper panel, with the modelled stratigraphy, to the lower panel with rheological profiles, indicate the area where the crustal rheology is adequately constrained (see also text). Abbreviations: *MTS* = Marl–Turbidite stage; *TS* = Transitional stage; *CPS* = Carbonate Platform stage; *RBS* = Red Bed stage; *T* = tension; *C* = compression; *QTZ* = quartz; *DIOR* = diorite. Numbers in the stratigraphic profiles mark ages of underlying horizons.

As shown by Figure 14, the pre-orogenic rheological structures predicted by the models are characterized by two minima in the strength profiles along the complete N–S traverse. The minima in the strength profiles represent zones that, upon application of lithospheric stress, could act as decollement levels. In the South Subbetic, the models predict a depth for these levels of 12–15 km and 22–25 km, respectively. To the north, the depth of the minima changes and lateral strength varies strongly. The error margins in these values are dependent on the error margins in the amount of stretching of the crust and are in the order of a few hundreds of metres. For other rheological properties of the upper and lower crust, the strength changes, but the locations of the minima remain the same.

As pointed out by Ord and Hobbs (1989), variations in lithospheric yield strength in a compressive setting also lead to the formation of ramp faults. The models thus predict ramp structures in the Iberian lithosphere in the Late Cretaceous. In the models with a 23-km initial EET, the decollements dip to the south. The models with a 30-km initial EET predict a variably dipping fault. At present no accurate information on the major decollement levels is available. However, future seismic reflection studies should allow to discriminate between the different model predictions.

The crucial aspect of the model predictions of the ramp structures is the important temporal and spatial variation of the rheological properties caused by stretching since Triassic times. Figure 14 shows a synthetic palaeo-rheological profile for the Iberian crust in the Triassic. For the calculations we adopt the temperature profile of the modelled elastic plate before stretching. The temperature profile is calculated starting from a depth of 100 km for the 1300° isotherm in undisturbed lithosphere. The lithospheric strength profile has been constructed with a three-layer model of the lithosphere consisting of a 16-km wet quartz layer, underlain by a 16 km wet diorite layer, and a 68-km basal olivine layer, respectively.

From Figure 14, it appears that the lower lithospheric strength increases in the areas that were affected by stretching. Stretching causes replacement of weak crust by strong mantle with an olivine rheology, leading to the formation of a dipping decollement level at the base of the crust. Furthermore, crustal domains of ductile deformation in the Triassic became brittle in the Jurassic–Cretaceous and vice versa. Since the generation and development of stress and strain patterns in the lithosphere are controlled by its rheological properties, it is obvious that Late Cretaceous and Tertiary basin development on the Iberian margin has strongly been affected by the preceding rifting stage.

As pointed out earlier, the predicted post-rift EET profiles are characterized by values of 22 to 14 km. In contrast, gravity and flexural foreland basin modelling of the Betics (Van der Beek and Cloetingh, 1992-this volume) yields estimates of the EET at present, which are less than 10 km. Since the Tertiary orogeny is associated with thickening of the lithosphere, the relatively small

TABLE 1

Rheological parameters used in the models to calculate depth-dependent rheological profiles of the Late Cretaceous Iberian margin (after Carter and Tsenn, 1987)

Layer	Mineralogy	Young's modulus (GPa)	Density ($g\ cm^{-3}$)	Creep parameters					
				N	Q_w ($kJ\ mol^{-1}$)	$\dot{\epsilon}_{01}$ ($Pa^{-N}\ s^{-1}$)	Q_D ($kJ\ mol^{-1}$)	σ_P (GPa)	$\dot{\epsilon}_{02}$ (s^{-1})
Upper crust	quartzite	25	2.6	1.9	172	$1.26\ 10^{-13}$			
Lower crust	diorite	35	2.9	2.4	212	$1.26\ 10^{-16}$			
Lithospheric mantle	olivine	70	3.3	3.0	510	$7.00\ 10^{-14}$	535	8.5	$5.7\ 10^{11}$

value of the present-day EET must be the result of another subsequent rifting phase. Geological studies and P–T–t modelling of the Internal zone provide strong evidence for heating as a consequence of thinning in the southern part of the Betics during the Late Oligocene–Aquitanian (García-Dueñas et al., 1986; Van Wees et al., 1992-this volume), while Kenter et al. (1990) provide sedimentological evidence for tensile tectonics in the Prebetic during the Aquitanian. As pointed out by several authors (Mauffret et al., 1984; Rehault et al., 1984; Burrus et al., 1987; Banda et al., 1992), Late Oligocene–Miocene rifting occurred in the Alboran Sea, the Valencia trough and the Gulf of Lions, which indicates that Miocene thinning of the southern part of the Iberian plate is related to regional tensile tectonics in the Mediterranean. Contemporaneous extension in the West European rift structure—from the Rhine Graben to the Rhône Graben (Illies, 1977; Bergerat, 1987)—suggests that rifting of the western Mediterranean should be regarded on an even larger scale. It thus appears that both the Mesozoic and Miocene rifting phases of the Betics reflect large-scale extensional tectonic processes in the Atlantic and European rift system evolution, respectively.

Conclusions

Quantitative subsidence analysis demonstrates that the Mesozoic history of the Iberian southern margin can be explained with multiple stretching phases at 220 Ma, 202 Ma and 160 Ma. The timing and duration of the stretching phases seems to be related to tensile regimes in the Atlantic region. Models of the stratigraphy of a segment of the Iberian rifted margin predict a N–S decrease of the EET of the lithosphere over a short distance, which implies a ramp-like structure of the lithosphere at the margin. The existence of such a structure is consistent with structural and stratigraphic data. As a consequence, the dynamics of the lithosphere during the Tertiary orogenic stage is significantly controlled by the pre-orogenic Mesozoic rifting stage.

The numerical models predict the location of major crustal weakness zones with ductile shear at depths of 9–12 km and 18–24 km. Deformation during Tertiary collisional tectonics will be concentrated in these zones of weakness, of which the depths are controlled by the preceding stretching phases. Our modelling has demonstrated that both the location and type of deformation in the orogenic stage of the Betics are to a large extent determined by the pre-orogenic phase of the tectonic evolution of the Betic margin.

Comparison of values for the present day EET with EET estimates inferred from the pre-orogenic rheology suggests that the mechanical properties of the lithosphere underlying the Betic margin were strongly affected by regional tensile tectonic processes in Oligocene–Neogene times, that also involved the Valencia trough, the Alboran Sea and the West European rift system. As such, quantitative analysis of the Betic margin provides important constraints for the dynamics of lithospheric extension on a larger scale.

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